3. Accuracy Requirements

Anthony DelGenio, NASA Goddard Institute for Space Studies

Satellite and surface measurements, if they are to serve as a climate monitoring system, must be accurate enough to permit detection of changes of climate parameters on decadal time scales. The accuracy requirements are difficult to define *a priori* since they depend on unknown future changes of climate forcings and feedbacks. As a framework for evaluation of candidate Climsat instruments and orbits, we estimate the accuracies that would be needed to measure changes expected over two decades based on theoretical considerations including GCM simulations and on observational evidence in cases where data are available for rates of change.

One major climate forcing known with reasonable accuracy is that caused by the anthropogenic homogeneously mixed greenhouse gases (CO_2, CFCs, CH_4 and N_2O). Their net forcing since the industrial revolution began is about 2 W/m^2 (Fig. 2.2), and it is presently increasing at a rate of about 1 W/m^2 per 20 years (Hansen and Lacis, 1990). Thus for a competing forcing or feedback to be important, it needs to be of the order of 0.25 W/m^2 or larger on this time scale.

The significance of most climate feedbacks depends on their sensitivity to temperature change. Therefore we begin with an estimate of decadal temperature change. Figure 3.1 shows the transient temperature trends simulated by the GISS GCM when subjected to various scenarios of trace gas concentration increases (Hansen *et al.*, 1988). Scenario B, which represents the most plausible near-term emission rates and includes intermittent forcing by volcanic aerosols, yields a global mean surface air temperature increase $\Delta Ts = 0.7°C$ over the time period 1995-2015. This is consistent with the IPCC projection of about 0.3°C/decade global warming (IPCC, 1990). Several of our estimates below are based on this assumed rate of warming.

Climate Forcings

Ozone. Ozone changes have the potential to be a major climate forcing, for which rates of change can be estimated from recent observations. Change of total column ozone during the 1980s was monitored by the TOMS satellite instrument (Fig. 3.2; Stolarski *et al.*, 1991). But, as indicated

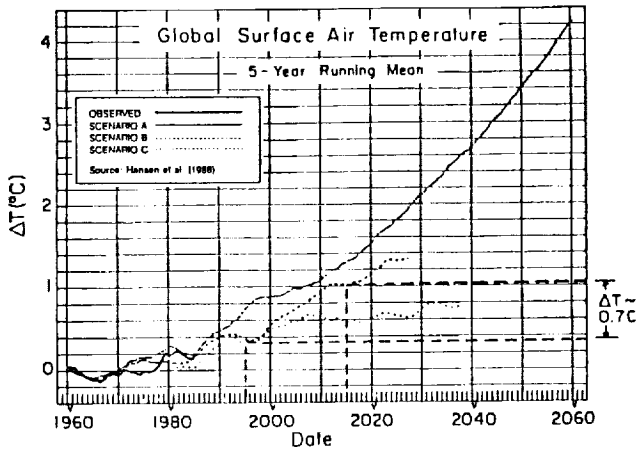

Fig. 3.1. Simulated global temperature change in climate simulations of Hansen *et al.* (1988). Rate of warming in the most realistic scenario (B), about 0.3°C/decade, is typical of other GCM and IPCC (1992) estimates.

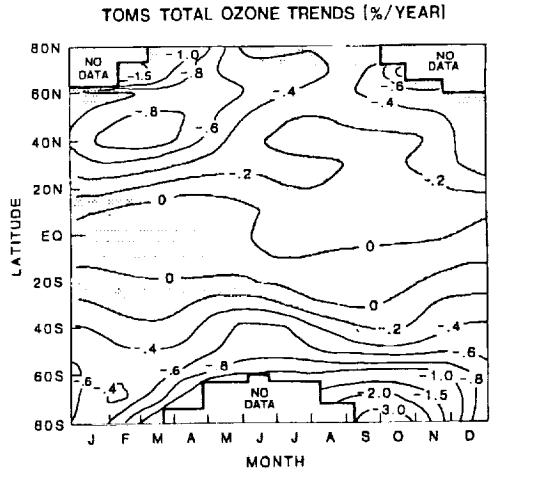

Fig. 3.2. Satellite (TOMS) observations indicate significant rate of total column ozone change during the 1980s (Stolarski *et al.*, 1991).

above (Figs. 2.2 and 2.3), the climate forcing due to the ozone change is entirely dependent on the vertical distribution of the ozone change. Data from a few mid-latitude ground stations suggest that the largest changes in the 1970s were near the tropopause (Fig. 2.4), and SAGE data for the 1980s suggest a qualitatively similar conclusion (McCormick *et al.*, 1992). The climate forcing by ozone depends mainly on the temperature of the ozone; as a result, it is required that the altitude of any significant ozone change be known within about 2 km in the troposphere and 5 km in the stratosphere. The magnitude of ozone change required to be significant is least at the tropopause, where changes of a few percent per decade are important, and increases toward both higher and lower altitudes.

Stratospheric water vapor. Doubling of stratospheric water vapor has been calculated to lead to a surface warming of the order of 1°C (Wang *et al.*, 1976), corresponding to a forcing of the order of 1 W/m^2. Thus, if the long-term change of stratospheric water vapor is monitored to a precision of 10 percent, its climate forcing can be defined very accurately.

Tropospheric aerosols. Tropospheric aerosols are thought to contribute substantially to climate forcing, but the magnitude of their impact is highly uncertain due to an absence of adequate global observations. Both anthropogenic and biogenic aerosols have received attention for their possible roles in climate change. Anthropogenic SO_2 emissions have probably at least doubled the sulfate aerosol concentration of the atmosphere over the past century relative to the background natural concentration (Fig. 3.3; Charlson *et al.*, 1992). Global increases of 10-20%, and regional increases of 50% or more, over a 20-year period are plausible. Such global aerosol changes could cause a direct aerosol forcing conceivably as large as 0.5 W/m^2, depending on the aerosol single scatter albedo, which would be comparable in magnitude to the expected climate forcing by anthropogenic greenhouse gases. Significant climate forcing from smoke due to biomass burning must also be considered (Penner *et al.*, 1992).

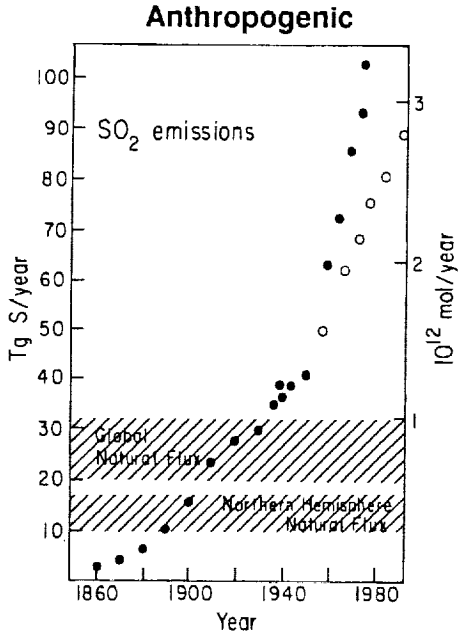

Fig. 3.3. Anthropogenic SO_2 emissions have probably at least doubled the sulfate aerosol concentration as evidenced by the above estimates for changes of SO_2 emissions. Open and filled circles represent data from two different sources (Charlson *et al.*, 1992).

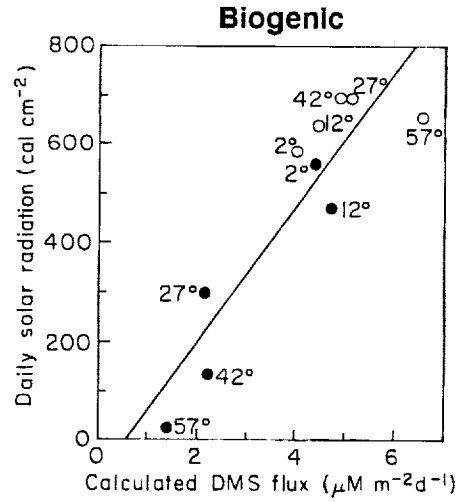

Fig. 3.4. Biogenic DMS emissions are sensitive to solar irradiance as evidenced by observed correlations between surface insolation and oceanic DMS flux (Bates *et al.*, 1987).

Biogenic emissions of dimethylsulphide (DMS) from the ocean appear to be sensitive to surface solar irradiance (Fig. 3.4; Bates *et al.*, 1987). Given the magnitude of measured solar luminosity variations (much less than one percent; Willson and Hudson, 1988) and the small variations in cloud cover and optical properties simulated by climate change models (Schlesinger and Mitchell, 1987), associated aerosol changes would be limited to a few percent over 20 years, much less than the anthropogenic component (Foley *et al.*, 1991). However, DMS emissions may also depend on other variable climate parameters, e.g., surface wind speed, in ways not currently documented. Climate forcing by a given aerosol optical depth is greater over the lower albedo ocean than over land; a change of global ocean aerosol mean optical depth of 0.01 is climatically significant (global forcing ~0.25 W/m^2). This change is an order of magnitude smaller than the accuracy or precision attainable with present satellite data.

The climate forcing by tropospheric aerosols depends on the aerosol optical depth, refractive index and size distribution, i.e., it is necessary to determine the aerosol microphysical properties (Patterson *et al.*, 1977; D'Almeida, 1987; Fouquart *et al.*, 1987; Tanre *et al.*, 1988; Leaitch and Isaac, 1991). A crucial parameter, which is very difficult to measure, is the aerosol single scatter albedo. One approach would be to infer the single scatter albedo by measuring the change of reflectance and aerosol optical depth together. The single scatter albedo needs to be known to an accuracy 0.02-0.03, which requires precision of the reflectance of the order of 0.01. Attainment of adequate knowledge of aerosol properties will require the combination of global satellite measurements supplemented by surface and in situ measurements for ground truth, as well as three-dimensional aerosol modeling.

Another major issue related to tropospheric aerosols is the changes which they may induce in cloud cover and cloud reflectivity. As an essential requirement for quantifying this climate forcing, the geographical distribution of aerosol microphysical properties must be monitored along with the cloud optical properties. The accuracy requirements for the measurements of cloud properties are described below.

Stratospheric aerosols. The climate forcing by stratospheric aerosols depends mainly on the visible optical depth of the aerosol layer, and secondarily on the aerosol size (Lacis *et al.*, 1992). Unlike the situation for the tropospheric aerosols, the forcing is practically independent of the amount of absorption by the aerosols (Lacis *et al.*, 1992). Addition of a visible optical depth of 0.15 causes a forcing of about 4 W/m^2, approximately the same as that for doubled CO_2, but in the opposite sense. Thus a significant climate forcing, 0.25 W/m^2, is caused by an optical depth of 0.01, which defines the required measurement accuracy. The effective radius of the aerosol size distribution needs to be known within about 50 percent.

Solar irradiance. A solar irradiance change of 2 percent, if spectrally flat, causes a climate forcing of 4-5 W/m^2, roughly equivalent to doubled CO_2. Thus a significant climate forcing would be produced by a solar irradiance change of about 0.1 percent, which defines the accuracy requirement for the integrated solar irradiance. However, climate forcing can also be caused by a change of the spectral distribution of the incoming radiation. The accuracy requirements are difficult to specify, because a change of the spectrum affects not only the amount and location of absorbed solar energy, but also may alter atmospheric composition, for example, ozone. The accuracies expected for the two spectral instruments on UARS, which monitor the sun in the ultraviolet region where the principal changes are known to occur, are probably sufficient, but measurements need to be extended to decadal time scales.

Surface reflectivity. A mean land reflectivity change of 0.1 is required to yield a forcing equivalent in magnitude to that for doubled CO_2 (Hansen *et al.*, 1988). Thus a significant global climate forcing (0.25 W/m^2) could result from a long-term mean surface reflectivity change of about 0.006. Since the mean land surface reflectivity is about 0.2, the long-term precision needed for surface reflectivity monitoring is about 2 percent.

Climate Feedbacks

Water vapor. The single largest positive feedback in GCM estimates of climate sensitivity is due to water vapor, the water vapor concentration increasing as climate warms. Climate models with a variety of approaches to the parameterizations of moist convection and stratiform clouds agree that *relative* humidity changes in a warming climate will be small, of the order of a few percent (Cess *et al.*, 1990; DelGenio *et al.*, 1991; Fig. 3.5). This implies large changes of *specific* humidity, i.e., water vapor concentration. Like ozone and clouds, though, the vertical distribution of the change is also important (Arking, 1993). Indeed, Lindzen (1990) has speculated that changes in moist convection in a warming climate could actually dry the upper troposphere enough to eliminate or reverse the water vapor feedback. Although, as discussed in Section 2, a broad range of scientific evidence argues against the extreme proposition of Lindzen, this does not reduce the need to better quantify the nature of the water vapor feedback by means of long-term monitoring of the change of the water vapor profile.

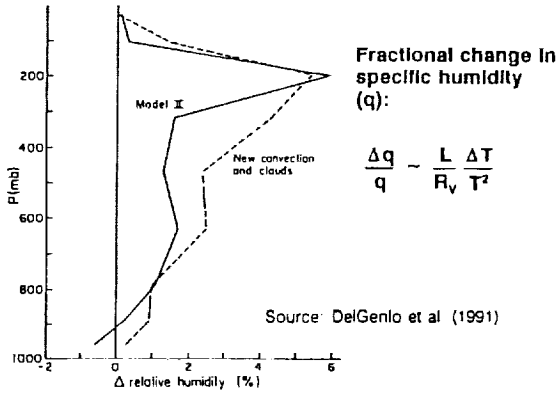

Fig. 3.5. Climate models with different approaches to the parameterization of convection and large-scale clouds yield similar responses of the water vapor profile to a change in surface temperature (DelGenio *et al.*, 1991).

If relative humidity changes are small, the Clausius-Clapeyron equation of thermodynamics can be used to estimate the change in specific humidity q from the change in saturation vapor pressure. The fractional change is $\Delta q/q \approx L\Delta T/(R_v T^2)$, where L is the latent heat of condensation, R_v the gas constant for water vapor, and T the temperature. For an assumed 0.7°C warming over 20 years, water vapor concentration would be expected to increase by about 4% (0.75 g/kg) near the surface and about 10% (0.001 g/kg) near the tropopause. Such a change of the water vapor profile, with everything else held fixed, would alter the net radiative flux at the tropopause or the top of the atmosphere by about 0.5-1.0 W/m^2, as indicated in Table 3.1.

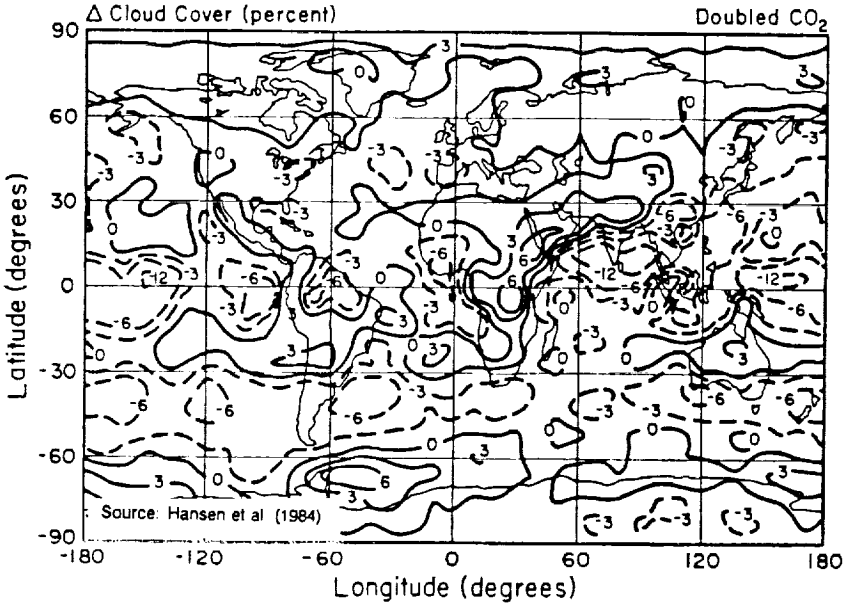

Fig. 3.6. Cloud cover changes in a doubled CO_2 experiment with the GISS GCM (Hansen *et al.*, 1984).

Cloud cover. There is no fundamental understanding of whether cloud cover should increase or decrease as climate warms, since the change depends on the subtle balance between the competing effects of moisture and temperature changes. Nonetheless, virtually all of the available GCMs which have been subjected to either doubled CO_2 (cf., Schlesinger and Mitchell, 1987) or a prescribed sea surface temperature anomaly (Cess *et al.*, 1990) predict that total cloud cover will slightly decrease as climate warms. The resulting cloud feedback on temperature is difficult to predict, because changes may be different for various cloud types, solar zenith angles, and underlying surface albedos. Simulated regional cloud changes are much larger than the global mean variation and may be of either sign (Fig. 3.6). Current GCMs suggest that over a 20-year period, global cloud amount may change by a fraction of a percent, with increases or decreases of 2-5% in different locations.

Ground-based observations of cloud cover (Henderson-Sellers, 1986, 1989; Karl and Steurer, 1990) suggest substantially larger variations over the past several decades, but the uncertainty in these observations is difficult to quantify. Satellite cloud observations (Rossow and Schiffer, 1991) show interannual global cloud changes of the order of a percent. A typical radiative flux change at the top of the atmosphere for a cloud cover change of 0.01 is 0.25 W/m^2 (Table 3.1) which is another indication of the cloud cover accuracy desired of observations.

Cloud height. All presently available GCMs predict that the mean altitude of cloud tops will rise in a warmer climate Fig. 3.7). There are several physical processes which influence the vertical distribution of clouds. In the tropics, for example, the dominant mechanism is deep moist convection, which supplies water vapor and ice for the formation of upper troposphere cirrus clouds and vents boundary layer water vapor that might otherwise form low-level stratus clouds. A simple estimate of the competing effects of boundary layer humidity and tropospheric lapse rate on convective stability and penetration depth (DelGenio, 1993) suggests that the altitude of tropical cirrus could rise by about 0.3 km (10-15 mb) in 20 years, given a 0.7°C surface warming. In midlatitudes, cloud height variations may be controlled additionally by changes in the strength and vertical scale of baroclinic waves. The change of cloud top level (and thus temperature) required to cause a change of 0.25 W/m^2 of the net radiative flux at the top of the atmosphere, evertyhing else held constant, is about 5 mb (Table 4.1).

Cloud optical thickness. Cloud optical thickness (τ) may be affected both by natural (i.e., thermodynamic and dynamic) and anthropogenic influences. Somerville and Remer (1984) used aircraft liquid water observations over the former Soviet Union to argue that low cloud optical thickness should increase by 4-5% per degree of temperature change. Theoretical arguments based on condensation in a lifted air parcel yield a similar result (Betts and Harshvardhan, 1987). Satellite data for the current climate confirm this finding over cold land areas, but suggest that at warm temperatures and especially over oceans, the optical thickness of low clouds may instead decrease with temperature, by as much as 10% per degree of warming (Tselioudis *et al.*, 1992). Thus, over 20 years with 0.7°C of surface

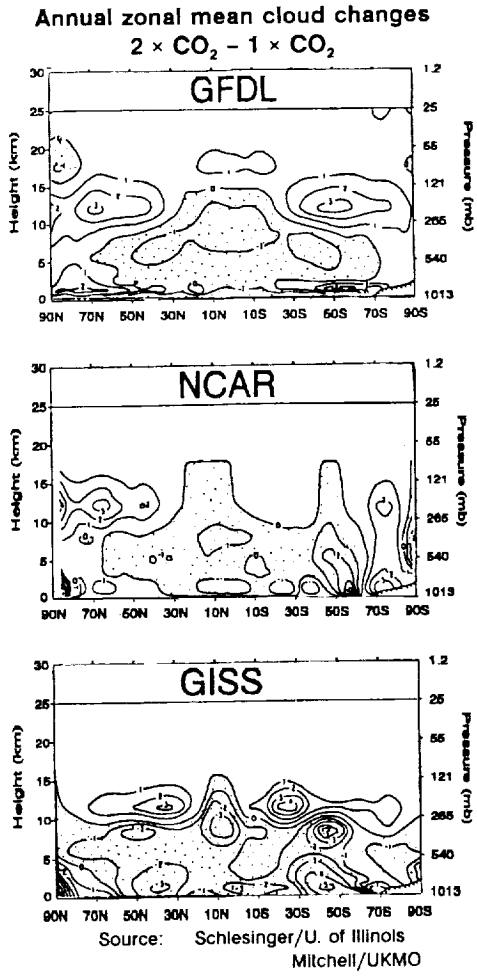

Fig. 3.7. Annual zonal mean cloud changes in equilibrium doubled CO_2 experiments with three GCMs (Schlesinger and Mitchell, 1987).

warming, a typical low cloud with $\tau = 10$ might experience an optical thickness change of about $\pm$ 0.5. If thin cirrus ($\tau \approx 1$) have similar temperature dependence, as suggested by the observations of Platt and Harshvardhan (1988), a much smaller change ($|\tau| < 0.1$) would be projected for that cloud type.

Another mechanism for changing cloud optical thickness is increasing aerosols, which may provide additional cloud condensation nuclei (CCN) for cloud droplet formation, especially over the remote oceans where droplet formation is most limited by the availability of nucleation sites. The effects are twofold. Increasing aerosols may translate into increasing droplet concentration and therefore decreasing droplet size, directly affecting cloud reflectivity (Twomey *et al.*, 1984). For example, if cloud liquid water content remains constant and if the fractional increase in droplet concentration is 70% as large as the increase in aerosol concentration (Kaufman *et al.*, 1991), the 50% regional increase in anthropogenic aerosols discussed above would imply a low cloud optical thickness increase of $\Delta\tau \approx 1$. Furthermore, smaller droplets precipitate less readily, suppressing an important sink of cloud water and thus further increasing cloud optical thickness (Albrecht, 1989). Where these effects occur, relative to the thermodynamic/dynamic controls mentioned previously, will influence the magnitude of the optical thickness changes. Quantitative analysis of the role of cloud optical thickness changes will require both global monitoring and in situ process studies.

Cloud particle size. Changes of cloud particle size are intimately involved in most mechanisms for change of cloud optical thickness. Although it is the change of optical thickness which is the immediate "cause" of a change of radiative balance, and thus of the climate forcing or feedback, it is important to measure the change of cloud microphysics. It is only with such knowledge that we are likely to obtain an understanding of the causes of any long-term changes of cloud radiative properties which affect climate sensitivity.

Cloud particle size changes are the result of the competing influences of changes in cloud water content and droplet concentration. In the absence of aerosol impacts, and ignoring sinks of cloud water such as precipitation and entrainment of clear air, the temperature dependence of adiabatic liquid water content (Betts and Harshvardhan, 1987) implies an increase in effective radius of only about 0.1 μm over 20 years. On the other hand, for a 50% regional increase in tropospheric aerosols, effective radius would be expected to decrease by approximately 1 μm. Size changes that large have been detected in ship tracks (Radke *et al.*, 1989) and between land and ocean clouds and northern and southern hemisphere clouds (Han, 1992). Determination of the climatic significance of this mechanism for cloud particle size change requires long-term global observations.

Radiative Impacts

The plausible changes over a 20 year period of the different climate forcing and feedback parameters discussed above can be readily converted to an approximate radiative flux change at the top of the atmosphere, all other factors being held fixed. The results of such computations are shown in the third column of Table 3.1, based on a radiative model employing the global datasets of the ISCCP project. The flux changes range from a few tenths of a W/m^2 to several W/m^2. These fluxes define the minimum accuracies that would be required for a global monitoring system. It is apparent that many of these fluxes are comparable in magnitude to the approximate 1 W/m^2 climate forcing which is anticipated to occur in the next 20 years due to continued increases of the homogeneously mixed greenhouse gases, CO_2, CFCs, CH_4 and N_2O (IPCC, 1992; Hansen *et al.*, 1988; Ramanathan *et al.*, 1985). The regional cloud changes are expected to be reduced on global average.

Ideally, a monitoring system for climate forcings and feedbacks would be capable not only of detecting changes of the magnitudes indicated in the first column of Table 3.1, but would measure any changes capable of yielding a significant forcing or feedback. We define a significant long-term global mean flux change as 0.25 W/m^2 or greater, based on the 1 W/m^2 forcing due to anticipated

TABLE 3.1. **Effect of anticipated parameter changes on radiative balance. Summary of anticipated or plausible changes of radiative quantities over a 20 year period (second column) as discussed in the text. The corresponding change in the net radiative flux at the top of the atmosphere is given in the third column, as estimated with a radiative model employing the global datasets of the ISCCP project.**

Forcing or Feedback	Anticipated Change of Quantity in 20 Years	Corresponding Δ Flux at TOA (W/m^2)
Ozone	$\frac{\Delta O_3}{O_3}$ = several percent or more	Latitude and Height-dependent
Tropospheric aerosol	$\Delta \tau$ = 0.04	−1.0
Stratospheric H_2O	$\frac{\Delta q}{q}$ = 0.3	+0.3
Surface albedo	ΔA_g = 0.01 (land)	−0.4
Tropospheric H_2O		
upper	$\frac{\Delta q}{q}$ = .10	+1.1
lower	$\frac{\Delta q}{q}$ = .04	+0.5
Cloud cover		
cirrus	Δ C = 0.03 (regional)	+2.0
stratus	Δ C = 0.03 (regional)	-3.0
Cloud top pressure	Δ p = −12 mb	+0.6
Cloud optical thickness		
cirrus	$\Delta \tau$ = 0.1 (regional)	+1.4
stratus	$\Delta \tau$ = 1 (regional)	−3.8
Cloud particle size	Δ r = −1 μm (regional)	−1.4

increases of greenhouse gases in the next 20 years. The constituent changes required to yield such flux changes are considered in Section 7 (Table 7.4). Many of these physical parameter changes are quite small. Nevertheless, we find that the potential exists for long-term monitoring of the climate forcings and feedbacks to precisions close to or exceeding even these more difficult requirements.